Sedimentology and carbon isotope stratigraphy of Lower–Middle Ordovician successions of Slemmestad (Oslo-Asker, Norway) and Brunflo (Jämtland, Sweden)

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Dissertations in Geology at Lund University,
Master's thesis, no 428
(45 hp/ECTS credits)
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Master’s thesis
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Department of Geology
Lund University
2015
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Cover Picture: The Hagastrand Member of the Tøyen Shale Formation at Bjerkåsholmen, Oslo-Asker.
Photo: Olof Peterffy
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Péterffy, O., 2015: Sedimentology and carbon isotope stratigraphy of Lower–Middle Ordovician successions of Slemmestad (Oslo-Asker, Norway) and Brunflo (Jämtland, Sweden). Dissertations in Geology at Lund University, No. 428, 41 pp. 45 hp (45 ECTS credits).

Abstract: Lower through Middle Ordovician strata are described with respect to sedimentology and carbon isotope stratigraphy from two areas, viz. two outcrops at Slemmestad in the Oslo-Asker district of southern Norway and a core section from the Brunflo area in Jämtland, central Sweden. The strata at these locations are compared and further correlation is done with the Tingskullen core section from Öland, south-eastern Sweden. Regionally important hardground complexes and beds such as ‘Blommiga bladet’ and ‘Blodläget’ and the ‘Volkhov-Kunda boundary bed’ are recognised in Jämtland and Slemmestad for the first time. The presented δ13C carb-values from Slemmestad indicate subsequent diagenetic alterations, which hampers regional correlation of the Huk Formation. δ13Corg from the Tøyen Shale Formation at this locality gives a more reliable signal and the BFICE has been tentatively recognised as 3.5‰ positive excursion in the uppermost part of the Hagastrand Member, although a higher data resolution would be needed to confirm it. The δ13C carb datasets from Brunflo and Tingskullen are both reliable and of high resolution. Several of the minor excursions and trends that characterise the generally stable carbon isotope record of the Lower and Middle Ordovician have been recognised, starting with the LTNICE, the BFICE, the Floian-Darriwilian rise and the BDNICE. A fast shift in the carbon isotope data is correlated to the ‘Täljsten’ interval which resents a regional biotic crisis. A negative excursion precedes the rising limb of the MDICE, which is clearly expressed (1.4‰) in the upper member of the Holen and Segerstad limestones in Brunflo. The correlation shows that the sedimentation rate was considerably higher in Slemmestad than in coeval strata in Jämtland and Öland during the Tremadocian and Floian; the Tremadocian is e.g. represented by almost 13 m of strata in Oslo as compared to the 2.5 m of coeval strata in Jämtland. The difference in sedimentation rate levelled out during the Dapingian and Darriwilian stages as a response to higher sea level at the time of deposition.

Keywords: Ordovician, carbon isotope stratigraphy, Baltoscandia, MDICE, ’Blommiga bladet’, ’Blodläget’

Supervisors: Mikael Calner, Oliver Lehnert

Subject: Bedrock Geology

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Sedimentologi och kolisotopstratigrafi avseende under- till mellanordoviciska avlagringar i Slemmestad (Oslo-Asker, Norge) och Brunflo (Jämtland, Sverige)

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Péterffy, O., 2015: Sedimentologi och kolisotopstratigrafi avseende under- till mellanordoviciska avlagringar i Slemmestad (Oslo-Asker, Norge) och Brunflo (Jämtland, Sverige). Examensarbeten i geologi vid Lunds universitet, Nr. 428, 41 sid. 45 hp.

Sammanfattning: Under- till mellanordoviciska avlagringar beskrivs med avseende på sedimentologi och kolisotopstratigrafi från två blottningar vid Slemmestad i Oslo-Askerdistriktet i södra Norge samt en borrkärnesektion från Brunfloodistriktet i Jämtland i Mittsverigeregionen. Avlagringarna från dessa platser jämfördes och korreleras med en borrkärnesektion från Tingskullen på Öland i sydöstra Sverige. Regionalt viktiga hårdbottenkomplex och bäddar, såsom Blommiga bladet, Blodläget och Volkohov-Kunda gränsbädden har påvisats i Jämtland och Slemmestad för första gången. De erhållna δ¹³C_carb-värdena från Slemmestad indikerar sentida påverkan av diagenetiska processer, något som inverkar negativt på Hukformationens regionala korrelationsmöjligheter. δ¹³C_org-data från Tøyenskifferformationen från den här lokalén ger en mer pålitlig signal och BFICE kan möjligtvis påvisas från lokalén som en 3,5 ‰ positiv excursjon i den övre delen av Hagastandledet, även om en högre datatätighet skulle behövas för att säkerställa att så verkligen är fallet. δ¹³C_carb-värdena från Brunflo och Tingskullen är både pålitliga och har hög uppläsning. Åtskilliga av de mindre excursjonerna och trenderna som känt emellan den generellt sett stabila kolisotoputvecklingen i under- och mellanordovicicum har påvisats, däribland LTNICE, BFICE, Floian-Darriwilianstigningen och BDNICE. En snabb skiftning i kolisotopdata korreleras med Täljstensintervallet som representerar en regional biotisk kris. En negativ excursjon föregår den stigande delen av MDICE, som är tydligt utvecklad (1,4 ‰) i det övre ledet av Holenkalisten samt Segerstadkalisten i Brunflo. Korrelationen visar att sedimentationstakten var avsevärt mycket högre i Slemmestad än i motsvarande avlagringar från Jämtland och Öland under Tremadocian- och Floianfaserna; Tremadocianfasen representeras t.ex. av uppemot 13 m mäktiga avlagringar i Oslo, att jämföra med de 2,5 m mäktiga liknande avlagringarna i Jämtland. Skillnaderna i sedimentationshastighet avtogs emellertid under Dapingian- och Darriwilianfaserna som ett svar på högre havsnivå vid bildningsställfället.

Nyckelord: Ordovicium, kolisotopstratigrafi, Baltoskandia, MDICE, Blommiga bladet, Blodläget
Handledare: Mikael Calner, Oliver Lehnert
Ämne: Berggrundsgeologi

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1 Introduction

The Ordovician was in many respects a unique period in Earth history (Jaanusson 1984). With continents widely separated (Cocks 2001) and extensive ocean floor spreading, the ocean waters were pushed up on the continents. The unprecedentedly high sea levels allowed vast continental areas to be flooded by shallow seas (Barnes 2004; Munnecke et al. 2010). Biodiversity boomed (Sepkoski 1981; Webby et al. 2004). This interval in the history of life is critical to understand but interpreting the depositional history, sea level changes and depths of ancient epeiric seas can be challenging. Modern analogues are lacking and intellectual problems arise from, for instance, condensation and reworking of sediments (e.g. Burchette & Wright 1992; Wright & Burchette 1998). To untangle these issues a broad set of tools must be employed, including traditional facies analysis but also geochemical data (Munnecke et al. 2010). Reliable correlations are vital to differentiate local from regional to global events. Over the last years, carbon isotope chemostratigraphy has become a useful tool for correlational purposes (e.g. Bergström et al. 2009; Saltzman & Thomas 2012).

The present study focuses on Lower and Middle Ordovician sedimentary successions on the western margin of Baltica, more precisely, on the outcrops near Slemmestad in the Oslo-Asker district, Norway and on a core from Brunflo, Jämtland (Fig. 1). Carbon isotope stratigraphy on whole rock analysis will be presented from both successions, as well as isotope data from organic carbon in the Tøyen Shale Formation at Slemmestad, the first organic carbon study from the Lower Ordovician of this part of Baltoscandia.

While the global record of the Upper Ordovician features several well-studied high amplitude carbon isotope excursions, the Lower and Middle Ordovician is characterised by gentle isotopic shifts and a more subdued generalised curve (Bergström et al. 2009). The most prominent excursion in the Lower and Middle Ordovician is the mid-Darriwilian isotopic carbon excursion (MIDICE; e.g. Meidla et al. 2004; Ainsaar et al. 2010, Schmitz et al. 2010), but lately, several smaller scale excursions have been described from Baltoscandia (Lehnert et al. 2014). Four of these are of interest of the present study, namely the LTNICE (Late Tremadocian Negative Isotopic Carbon Excursion), BFICE (Basal Floian Isotopic Carbon Excursion), which marks the onset of the Floian-Darriwilian rise), BDNICE (Basal Dapingian Negative Isotopic Carbon Excursion) as well as the LDNICE (Lower Darriwilian Negative Isotopic Carbon Excursion). All of these have been recognised in the data presented from the Brunflo #2 core.

While sea level changes in the shallow, central parts of the basin may give rise to sharp facies shifts, large parts of the successions may also be cut out by erosion. The marginal marine strata provide a more complete stratigraphic record, even though any facies shifts related to sea level changes may be subtle. The aim of the present study is therefore to correlate these marginal deposits to the more central parts of the Baltoscandian basin in order to unravel the history of sea level changes at the time of deposition. A core recovered from Tingskullen on the island of Öland, Sweden, with a high resolution carbon isotope stratigraphy (Calner et al. 2014) linked to a detailed conodont biozonation by Wu et al. (submitted) will be used for comparison.

2 Geological setting

Three main physical aspects influenced environment and sedimentation in Baltica in the early Ordovician. The first of these was tectonically induced, as the continent was moving northwards across the southern hemisphere to higher latitudes with a warmer climate (Cocks & Torsvik 2005). This allowed the deposition of cold water carbonates in the Baltoscandian basin which hitherto in the Cambrian and lowermost Ordovician had been dominated by siliciclastic sedimentation (Calner et al. 2013). Secondly, the continent had been subject to a prolonged period of erosion and epeiric quiescence in the Precambrian, a process which had left the continent essentially flat and peneplanised (Lidmar-Bergström 1993, 1995). This had major consequences in combination with the third factor, namely the extraordinary high sea level.

Although the eustatic sea level fell somewhat in the Middle Ordovician, sea levels had risen since the Late Precambrian, and the Early Ordovician recorded the second highest sea level of the entire Palaeozoic (Nielsen 2004; Haq & Shutter 2008). In combination with the low relief, this allowed large parts of the continental surface to become inundated by a shallow epeiric sea. The low relief in itself, in combination with the drowning of source areas, led to extreme sediment starvation. Average net accumulation rates were merely 1–9 mm/1000 years in Sweden and the East Baltic area at the time, whereas the Oslo area had slightly higher values of 3–12 mm/1000 years, as it was located in the distal foreland of the Caledonides (Nielsen 2004). As a consequence, the Baltoscandian Cambrian to Middle Ordovician successions are relatively thin (Calner et al. 2013) and the entire Ordovician in Norway is represented by merely 400 m of strata (Bockelie 1982). Fluctuations in the relative sea level gave rise to the cyclic pattern displayed in the strata, with shale representing deeper conditions and limestone more shallow environments (Egenhoff et al. 2010).

The depositional and ecological environment in the basin varied with e.g. depth and distance from the coast line. Based on similarities in lithology and corresponding ecologic and faunal zonation, Jaanusson (1972, 1982, 1995) distinguished four ‘Confacies Belts’ in the Baltoscandian basin (Fig. 1A). The Scanian and Oslo belts are the most offshore of these, whereas the North Estonian with its southern counter-
part, the Lithuanian Facies Belt represent the most nearshore conditions. The Central Baltoscandian Belt and its projection, the Livonian tongue, represent an intermediate position. The Oslo Facies Belt distinguishes itself in having a broad variation in lithofacies and faunal composition, likely reflecting a pronounced bottom topography (Pärnaste et al. 2013), with the Oslo-Asker area situated in the deepest parts of the Oslo Facies Belt (Fig. 2).

The 40-70 km wide and 115 km long Oslo Region (Oslofältet) preserves Lower Palaeozoic strata extending over an area of 10000 km² (Nakrem & Rasmussen 2013). The stratigraphic outline is presented in Figure 3. The Ordovician sedimentary successions of the Oslo Region have been studied by palaeontologists and sedimentologists for more than 150 years, with early contributions from e.g. Kjerulf (1857) and Brøgger (1882). Størmer (1953) outlined the regional lithostratigraphy and divided the region into different districts. The lithostratigraphy was subsequently thoroughly revised by Owen et al. (1990). The first general survey of the geochemistry of the strata was carried out by Bjørllykke (1974). Modern palaeontological studies include contributions by Ebbestad (1999; trilobites of the Bjørkåsholmen Formation), Nielsen (1995, trilobites of the Huk Formation), Hansen (2009, trilobites of the Elnes Formation), Rasmussen (1991, 2001; conodonts of the Huk Formation and coeval Stein Formation respectively) and Hoel (1999a,b; trilobites of the Tøyen Shale Formation).

The Oslo Region constituted a passive continental margin in the Cambrian and Early Ordovician, but with the incipient stages of the Caledonian Orogeny in the west during mid Darriwilian times (Elnes Formation) and onward, more siliciclastics were introduced into the basin (Bruton et al. 2010). At the peak of the orogeny in the Silurian, the successions were thrust south-eastwards and folded; as a result, the studied successions in the Oslo Region dip steeply. Commonly, the Cambrian-Ordovician Alum Shale acted as a decollément plane (Bruton et al. 2010). Extension in the Permian further disturbed the successions both by the intrusion of dolerite dykes, as well as block faulting (Bockelie 1982), but if the succession had not been faulted down, they would have been eroded away and not be preserved today (Hansen et al. 2011). In other words, even when the boundaries of the preserved basin are those of a Permian aulacogen, the Early Palaeozoic sediments were originally deposited in an epicratonic basin (Owen et al. 1990).

Much like the Norwegian Early Ordovician successions, the Jämtland counterparts were also formed in a shallow epicontinental sea at the western, passive mar-

Fig. 1. A: Map of Baltoscandia with Lower Palaeozoic strata indicated. Facies belts after Januusson (1995). B: Local map of the vicinity of Slemmestad. Field work was carried out at VEAS and Bjørkåsholmen C: Local map of Brunflo where the Brunflo #2 core was recovered.
gin of Baltica that subsequently were subjected to deformation due to the Caledonian Orogeny (e.g. Karis 1998). Even though the Brunflo area is considered to be part of the autochthonous sequences of Jämtland, a minor degree of thrusting and faulting of the strata took place (Löfgren 1978). The Jämtland successions have been divided into a platform and ramp facies, where the Brunflo area constitutes the former because of the predominance of carbonate-dominated successions (Karis 1998). Accordingly, the palaeosetting is slightly shallower than the Slemmestad counterparts, and corresponds to the Central Confacies Belt (Fig. 1A, Fig. 2; Jaanusson 1995, Karis 1998).

The biostratigraphy of the Lower and Middle Ordovician of Jämtland is not particularly well known, but it has partly been studied with respect to conodonts (Löfgren 1978, Sturkell 1991), trilobites (Tjernvik 1956, Larsson 1973) and agnostids (Ahlberg 1988). Several reviews and excursion guides of the area have been published, including Bruton & Williams (1982), Karis (1998) and Wickström (2007). The stratigraphy is outlined in Figure 4.

3 Material and methods

Studies were undertaken in two successions which formed at the western margin of Baltica, the first represented by two outcrops in the vicinity of Slemmestad in Oslo-Asker and the other one recorded in a core recovered from the Brunflo area of Jämtland. More emphasis was put on the two outcrops in the former area (Fig. 1B). At the first outcrop, the beach at the Bjerkåsholmen peninsula, strata ranging from the Tremadocian part of the Alum Shale to the Darriwilian Sverrotten Member of the Huk Formation are exposed. The Tøyen Shale Formation is only partly exposed. The succession is partly repeated at the second locality, a road cut at the Djuptrekkodden peninsula just 200 m to the north, also known as the VÆS section (Fig. 1B). The outcrop there includes the uppermost and tectonically deformed Tøyen Shale Formation, the entire Huk Formation as well as the basal part of the Elnes Formation.

The contemporaneous succession deposited at Brunflo, Jämtland (Fig.1C) was studied in the Brunflo #2 core which was drilled by the Geological Survey of Sweden (SGU) in 1970. The core section records autochthonous strata ranging from the Cambrian part of the Alum Shale Formation through a major part of the Lower-Middle Ordovician 'orthoceratite limestone', up to the Segerstad Limestone. The core, which is stored at the SGU office in Malå was sampled for carbon isotope chemostratigraphy by Rongchang Wu in early 2014 and further documented by the present author with respect to sedimentology in September 2014. These data was published in Wu et al. (in press) and are referred to below.

Field work was carried out at Slemmestad in May and October of 2014. At the later date the entire VÆS section had been thoroughly excavated, likely to reduce risk of slumping over the road. This has given new and relatively unweathered surfaces of the Huk Formation, but the Elnes Formation is now almost completely obscured at the locality. The sections were measured, logged and studied with respect to their macroscopical properties. The logs presented here are composite logs mainly based on the Bjerkåsholmen exposure, with the exception of the Sverrotten Member and the Elnes Formation, which were logged at the VÆS section. The former unit is substantially better exposed at this locality after the excavation and the latter was logged prior to the excavation.

The lithology and faunal composition were studied using a polarised light microscope from about twenty thin sections that were prepared from rock samples collected at Bjerkåsholmen and the VÆS section. Furthermore, several rock samples were polished and examined using a standard optical microscope and a hand lens. Due to time constraints, no thin sections were prepared from the Brunflo #2 core, as a result only the latter method was employed. The sedimentological findings were then compared to the carbon isotope data of Wu et al. (in press) sampled in the same core.

A total of 106 samples for studies of bulk δ¹³C_carb were collected from the limestone dominated formations at Slemmestad, i.e. the 10.6 m of strata that make up the Bjerkåsholmen and Huk formations as well as the exposed part of the Elnes Formation, yield-
Fig. 3. Litho- and biostratigraphy of the central Oslo Region. From Nakrem & Rasmussen (2013), modified from Bruton et al. (2010). The interval studied in this thesis is marked by the blue box.
an average sampling density of one sample per 0.1 m. The first two formations were sampled at the Bjerkåsholmen peninsula and the latter at the Djuptrekkodden road cut. Samples were obtained using a micro-drill on fresh rock surfaces. Cement and calcite veins were avoided, and only micritic lithologies were targeted. Carbonate powders were subsequently reacted with 100% phosphoric acid ($\text{H}_3\text{PO}_4$) at 70 °C with a Gasbench II connected to a ThermoFinnigan Five Plus mass spectrometer. All reported values are in parts per mil relative to Vienna Pee Dee Bel-lemnite (V-PDB) by assigning $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of $+1.95\%$ and $-2.20\%$ to the international standard NBS19 and $-46.6\%$ and $-26.7\%$ to the international standard LSVEC, respectively. Reproducibility and accuracy of carbon isotope analyses were monitored by replicate analysis of laboratory standards which were calibrated to NBS19 and LSVEC and were $±0.05\%$ ($±1$ std. dev.).

Thirty-three samples for organic $\delta^{13}\text{C}$ studies were recovered from the 16.3 m of marine shales comprising the Tøyen Shale Formation, giving an average sampling density of about one sample per 0.5 m. Each sample was washed and then ground to a fine powder. 10 % hydrochloric acid was added after which the samples were allowed to react for several hours, a procedure which was repeated for the carbonate-rich samples several times. The acid was then decanted after which the samples were thoroughly washed with distilled water. Carbon isotope analysis of organic carbon was performed with an elemental analyser (CE 1110) connected online to a ThermoFinnigan Delta V Plus mass spectrometer. All carbon isotope values are reported in the conventional $\delta$-notation in permil relative to V-PDB. Accuracy and reproducibility of the analyses was checked by replicate analyses of international or laboratory standards (USGS 40 and ERL 5). Reproducibility was better than $±0.05\%$ ($1\sigma$).

Analysis of elemental compositions was performed using scanning electron microscopy and EDX on glauconitic levels and levels with suspected pseudomorphs. Fresh rock samples were mounted in epoxy for easy handling and subsequently polished smooth using 1 micron diamond micropolish. The samples were carbon coated before analysis in a variable pressure Hitachi 3400N scanning electron microscope.

4 Stratigraphy

4.1 Slemmestad outcrops

The Bjerkåsholmen peninsula and the VEAS section on Djuptrekkodden near Slemmestad, Oslo-Asker, exposes strata ranging from the upper Cambrian Alum Shale Formation to the middle Darriwilian Elnes formations. The strata dip steeply to the north-west and the formalional boundaries are indicated in a detailed geological map (Fig. 5).

4.1.1 Alum Shale Formation

The Alum Shale Formation can be recognised over a vast area across the Baltoscandic platform, from Poland in the south to the East Baltic area in the east and even beneath the Caledonian nappes on the west coast of Norway (Gee 1980). It ranges stratigraphically from the middle Cambrian through the Tremadocian, its base being diachronous (Nielsen & Schovsbo 2007; Bruton et al. 2010 and references therein). The upper part of the shale unit in the Oslo Region is poorly fossiliferous, but has yielded specimens of the trilobite *Ceratopyge forficula* as well as the graptolites *Kiaeroegraptus kiaeri* and *Bryograptus ramosus* (Owen et al. 1990). Ebbestad (1999) assigns the top of the formation to the *Kiaeroegraptus supremus* graptolite Zone.
of the Varangu Regional substage. The topmost part of the formation was previously known as the *Ceratopyge* Shale (Owen et al. 1990).

The Alum Shale Formation is severely deformed, displays prominent folds and is furthermore largely eroded away at Bjerkåsholmen. A detailed study of the formation is beyond the scope of this thesis. The black and highly kerogenous shales are poor in carbonate beds and carbonate concretions (or ‘stinkstones’) at Slemmestad and in the Oslo area (Owen et al. 1990).

The lithology is fairly uniform, but the frequency of grey silt beds increases in the upper 0.6 m. A horizon with 2–3 cm long, black and ellipsoidal fine grained limestone concretions 0.06 m below the contact to the overlying Bjørkåsholmen Formation.

### 4.1.2 Bjørkåsholmen Formation

The 1.2 m thick Bjørkåsholmen Formation is widely distributed in the Oslo Region (e.g. Fjeldal 1966; Ebbestad 1999) and comprises micropartic and micritic limestones with two thick shale interbeds (Figs. 6–7). Original depositional structures are hard to discern due to pervasive recrystallisation and diagenetic alterations (Fig. 8). The formation was formerly known as the Ceratopyge Limestone, but was redefined as the Bjørkåsholmen Formation by Owen et al. (1990).

Analysis of the trilobite faunas performed by Ebbestad (1999) shows that the thin unit spans the *Apatokephalus serratus* Zone. The basal nodular bed contains the *Ceratopyge* fauna, i.e. the fauna associated with *Ceratopyge forcifica* and *Ceratopyge acicularis*. The top of the formation coincides with the top of the *Paltodus deltifer* conodont Biozone (Erdtmann & Paalits 1994).

The base of the formation is undulating and sharp, with a basal bed that constitutes wackestone nodules within a sparitic, recrystallised matrix (Fig. 1A,D). These pale grey nodules are well rounded but irregularly shaped, up to 0.2 m long, and some show signs of brittle deformation (Fig. 7A,B). Trilobite bioclasts dominate, but brachiopods and echinoderms also make up a substantial part of the fauna. Fine grained pyrite crystals (0.1 mm) are dispersed throughout the bed, but are more highly concentrated in the nodules.

The basal limestone bed is overlain by a 0.14 m thick dark grey shale bed that includes a layer of isolated black mudstone nodules. These nodules are dominated by the *Bienvillia angelini* olenid fauna and, as they can be traced throughout the Oslo region, the nodule layer acts as an important marker bed in the region (Ebbestad 1999).

The shale bed is succeeded by a 0.4 m massive set of six thin to medium thick limestone beds. This unit was named ‘the main limestone bed’ by Fjeldal (1966). The contact with the underlying shale is abrupt, and at the base black, well rounded limestone clasts occur again. These are macroscopically similar to those of the *Bienvillia angelini* bed but are here separated laterally by a 2–7 cm thick layer of fibrous crystalline calcite or orsten, resembling cone-in-cone structures (cf. Tucker 2001; Cobbold et al. 2013). The calcite fibres are conically arranged and crystal axes are oriented oblique to bedding. Calcite cone-in-cone...
Fig. 7. Photographs of Bjørkåsholmen and Tøyen formations at Djuptrekkodden. A: The 1.2 m thick Bjørkåsholmen Formation in its entirety. B: Nodules (pale) set in a recrystallized matrix (orange) in the basal limestone bed of Bjørkåsholmen Fm. The upper nodule appears brittlely deformed (at pen-tip). C: Two beds with white calcium carbonate crystal aggregates possibly representing pseudomorphs after gypsum, set in a recrystallized matrix. The grey middle bed is a wacke-packstone that has not undergone substantial recrystallisation. D: Possible gypsum pseudomorph in the upper part of Hagastrand Member with conspicuous swallow-tail morphology. E: Multicoloured horizon at the base of the informal Stemnestad Member (Erdtmann 1965), possibly correlating with the ‘Blommiga bladet’ hardground. It is approximated as the base of the Dapingian in the section F: Contact between Tøyen and Huk formations.
Fig. 8. Photomicrographs from the Björkåsholmen Formation. A: The basal bed shows trilobite wackestone nodules set in an almost thoroughly recrystallized matrix. A few trilobite grains have avoided recrystallisation. B: A bedding parallel laminated fabric is seen in the lower part. A sharp boundary divides this bed from the upper one which shows 1-2 cm long calcite aggregates which may represent gypsum pseudomorphs, set in a recrystallized matrix. Trilobite wackestone clast in the top. C: The uppermost bed of the Björkåsholmen Formation consists of crystal aggregates of uncertain affinity. Large opaque minerals are pyrite. Note cross-shaped pyrite crystal in the lower part, a shape indicative of replacement of gypsum. D: Close-up of (A), contact between recrystallized matrix and better-preserved nodule. E: Laminated lower part of (B). F: Close-up of upper part of (B). G: Close-up of (C). Note cross-shaped pyrite in upper right.
structures are commonly found in marine shales and are indicative of fluid overpressure during diagenesis (Cobbeld et al. 2013).

Two beds (Fig. 6) feature conspicuous calcite crystals in ovoid or spherical shapes (Figs. 7C, 8B,F). Both of these beds are underlain by laminated wackestones. The calcite crystals seem to have grown in an un lithified, plastically deformed mud (Fig. 8B) and they have previously been interpreted as possible pseudomorphs of gypsum (Fjelldal 1966). The matrix itself show a gradation, with micrite in the upper part, transitioning to microspar in the middle part of the bed and down into pseudospar surrounding the ovoid crystals (Scholle & Ulmer Scholle 2003, p. 270). Only few bioclasts occur in the sparitic matrix.

The ‘main limestone bed’ is capped by a thin bed of micritic packstone that does not show signs of recrystallisation. A ~0.2 m thick grey shale bed separates this unit from the topmost, ~0.4 m thick bedset of the formation. These beds are thoroughly recrystallised. The basal bed is a pale grey microsparitic mudstone with a few skeletal grains of mainly trilobites in its lower part. This horizon shows wavy bedding and varies in thickness from 0.1 to 0.3 metres.

The wavy bed is overlain by three thin and dark grey limestone bands, totalling 0.18 m in thickness. These beds are composed of sand sized, green aggregates in a dark grey, cryptocrystalline quartz matrix. While the beds are entirely devoid of bioclasts, pyrite concretions are common and these occur both with framboidal morphology as well as twinned crosses (Fig. 8C,G). The twinned crystal habit is very similar to gypsum crystals documented in Scholle & Ulmer Scholle (2003, p. 395), so the pyrite may well represent gypsum pseudomorphs. The green grains in these beds are irregular but rounded in shape, with radial cracks and have been interpreted as glauconite (e.g. Egenhoff et al. 2010). These grains are, however, not green in thin section (Fig. B10) and X-ray diffraction studies on glauconitic grains from Vestfossen and Bjerkåsholmen revealed only the presence of illite (Fjelldal 1966, Bjørlýkke 1974). As a consequence Bjørlýkke (1974) hypothesised that the grains were presumably glauconitic originally, but that they were diagenetically altered to illite. The chemical composition of these grains were analysed using EDX and the results are presented in Table 1. They have an average potassium content of 7.34%.

### Table 1. Elemental composition of glauconite in the Bjerkåsholmen Formation.

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#### 4.1.3 Tøyen Shale Formation

The Tøyen Shale Formation and its lithostratigraphic subdivision into three members was defined by Erdtmann (1965). The lowermost Hagastrand Member is characterised by grey shales and silty interbeds whereas the middle Galgeberg Member comprises dark, graptolitic shales. A third member, the Slemstad Member, was also defined by Erdtmann (1965) for the uppermost part of the formation where the shales are paler and have a higher carbonate content. This member was subsequently discarded by Owen et al. (1990) however, as they considered this division was based on faunal rather than lithological differences.

The Tøyen Shale Formation is poorly exposed in the Oslo Region, the original stratotype section at Tøyen, Oslo where it was defined by Erdtmann (1965) was only briefly exposed during the construction of a railway tunnel. Therefore, Owen et al. (1990) designated a neostotype at Engervik. These two publications, however, define the base of the Galgeberg Member differently. A level where graptolites appear defines the base of the Galgeberg Member according to Erdtmann (1965) whereas Owen et al. (1990) define the base of the Galgeberg Member where mudstone and grey shale interbeds cease. As Erdtmann based his division more on faunal rather than lithological basis, the definition by Owen et al. (1990) is preferred here.

The Tøyen Shale Formation is absent in the southern part of the Oslo Region, but can otherwise be recognised throughout the area (Owen et al. 1990). It also extends into Västergötland, Jämtland and Skåne in Sweden (e.g. volume by Bruten & Williams 1982). The unit is dominated by shales in the Oslo Region, but in Eiker-Sandsvær, the Hagastrand Member is instead developed as a condensed, glauconitic limestone. Hoel (1999a) documented a diverse fauna of trilobites belonging to the Megistaspis (Paramegistaspis) planilimbata Zone in this facies. The biostratigraphy of the shaly facies is less well known; no recent biostratigraphical data from the Tøyen Shale Formation in Norway is known to the author. Furthermore, the Hagastrand Member is only sparsely graptolitic and Erdtmann (1965) found no graptolites at all in the lowermost 3 m. In contrast, the black shales of the overlying Galgeberg Member are rich in graptolites. Coinciding with the base of the member are the first appearances of Tetraraptus approximatus and Tetraraptus phylograptoideus, thus marking the base of the Floian stage (Cooper & Sadler 2012).

A total of 16.3 m of the Tøyen Shale Formation is exposed at the Bjerkåsholmen peninsula; the lower middle and middle part of the Galgeberg Member is obscured by Quaternary deposits (Fig. 5). Judging from the near-vertical dip of the strata and the distance between outcrops, the total thickness is estimated at 28 m, i.e. 8 m thicker than was described from the type locality at Tøyen, Oslo, by Erdtmann (1965). Tectonic thickening of the succession cannot be excluded.
Fig. 9. Log of Töyen Shale formation at Djuptrekkodden in Slemestad. Most of the Galgeberg Member is obscured by quaternary deposits at the locality due to the weathering prone black shale lithology.
The greenish ‘glaucninite’ beds of the Bjørkåsholmen Formation end abruptly and is succeeded by the shales of the Hagastrand Member. Both the base and the top of the member are dominated by grey shales with calcareous mudstone interbeds whereas dark grey shales with black interbeds constitute the middle part (Fig. 9).

Rhythmically distributed, ~0.1 m thick beds, characterise the lowermost 4 metres of the formation. These are impure dolomitized mudstones, often containing large, up to 0.05 m long pyrite concretions arranged in bands.

About 1 m above the base, 5 mm large and often diamond shaped crystals and holes (from where the crystals have been weathered out) are common. They have been considered to represent pseudomorphs of gypsum, thus indicating evaporitic conditions (e.g. Erdtmann 1965). However, based on their shape and the high concentrations of barium in these beds, Antun (1967) and Bjørlykke (1974) suggested that they are pseudomorphs of barite (BaSO4). None of the presumed barite remains though, as it has been replaced by calcium carbonates, granular quartz and pyrite (Antun 1967). Therefore, the conclusion that these crystals originally represented barite (and not any other sulphate mineral) is by no means unambiguous.

Between ca 4 and 7 m above the base, the shales become more fissile than below, and black shales are interbedded in the grey shales. Silty interbeds are absent in this interval. Nonetheless, at 7 m pyrite aggregates (1-2 cm in diameter) can be observed, as well as holes that may either represent casts where the pyrite has been weathered out or possibly moulds of dissolved authigenic minerals.

An interval with grey shales with presumed pseudomorphs (Fig. 8D) and silty interbeds reappears at 8.45 m from the base of the formation, capped by a prominent 0.2 m thick mudstone bed. Above this bed, black and grey shale beds intercalations are observed and the lithology becomes progressively more black and fissile. Above a level of mudstone lenses at 12.3 m the strata are poorly exposed and covered by Quaternary deposits.

When comparing the observations of the present study with the log from Engervik by Owen et al. (1990), the base of the Galgeberg Member (including the informal, 1.3 m thick Slemmestad Member) at Bjørkåsholmen is inferred to be just above the level where the strata start to be obscured. In other words, the Galgeberg Member is not exposed in its entirety at Djuptrekkodden. The unexposed part of the member is estimated to be more than 12 m thick at the locality and 3.5 m of Galgeberg strata can be observed above the weathered out portion, just below the Huk Formation. Provided that the obscured strata are not folded or otherwise deformed, the thickness of the Galgeberg Member is up to 16 m thick.

The shales of Galgeberg Member above the obscured part are dark and fissile, but at 1.3 m below the base of the Huk Formation they culminate in a prominent pyrite impregnated horizon that shifts in bright colours ranging from blue-green to yellow to orange (Fig. 7E; see discussion for possible correlation). Above this level, the shales become progressively greyer and more carbonate rich (cf. Slemmestad Member as defined by Erdtmann, 1965).

### 4.1.4 Huk Formation

A sharp lithological change marks the transition from the Toyen Shale Formation to the limestones of the Huk Formation, which is ca 8.5 m thick in the study area. The formation is part of the so called ‘orthoceratite limestone’, which has a wide distribution in Baltoscandia. The lithology shows a tripartite development in the Oslo Region and was divided into the Hukodden, Lysaker and Svartodden members by Owen et al. (1990). The basal and topmost members feature dark grey, compact limestones whereas the middle member is characterised by less weathering resistant limestone-marl alternations (Fig. 10). The formation’s hypostatotype is located in Huk, Oslo-Asker.

#### 4.1.4.1 Hukodden Member

The 1.7 m thick Hukodden Member comprises compact, greenish grey wacke- to packstones. The contact to the underlying Toyen Shale Formation is distinct, a feature that is enhanced by weathering of the less resistant shales (Fig. 7F).

The basal beds show a wide lithological variation and call for a more detailed description. The basal 10 cm are made up of a thoroughly recrystallised mudstone with few remaining skeletal clasts. The lowermost centimetre displays a slight fissility, reflecting a certain amount of terrigenous material in the basal lime mudstone. Above this interval the shale content decreases rapidly, but the rock becomes progressively darker, culminating in a dark band 5 cm from the base. This layer likely reflects a higher content of organic material. Dark structures that may represent burrows extend upwards and downwards from this band, and in places the burrows can also be traced laterally. These traces have a flattened appearance, suggesting post-compressional lithification. Above this horizon the limestone shows the same recrystallisation as below, but is substantially paler.

A boundary, which appears at 0.1–0.13 cm from the base of the formation, separates the recrystallised rock from a dense trilobite and brachiopod packstone. The dark grey mineralised zone is 5 mm thick and has a sharp, well defined top but a gradual base. Laterally, the contact is undulating but smooth, and shows no signs of boring. The packstone bed above this zone is only 4 cm thick and is capped by a prominent hardground, 0.17 m above the base (Fig. 11A–B). The irregular surface is impregnated by millimetre-sized pyrite cubes and narrow cylindrical borings extending up to 3 cm below the surface. According to Rasmussen (1991, p. 267) and Rasmussen et al. (2013), this sur-
Fig. 10. Composite log of Huk Formation at Djuptrekkodden and the VEAS section with carbon isotope data from the limestones, displaying a diagenetic overprint of the strata. The Huk Formation shows a clear lithological differentiation between the three members; the Hukodden and Svartodden members are dominated by compact wacke-packstones whereas the Lysaker Member is characterized by rhythmic limestone-marl interbeds.
face represents a hiatus spanning from the middle Bal
toniodus navis to the basal Paroistodus originalis con-
odont biozones). Furthermore, it marks the boundary
between the Megistaspis polyphemus and Megistaspis
simon trilobite zones (Nielsen 1995; Fig. 10). This
makes it coeval with ‘Blödläget’, a hardground com-
plex developed as a bed of three hematite impregnated
surfaces on Öland, where it also marks the base of the
Parastoidus originalis Biozone (Lindström 1979; see
discussion).

Above this hardground, the lithology is dominated
by bioturbated wackestone-packstone. Dark blue grey
patches of packstone with grains oriented parallel to
the bedding are separated by centimetre-thick greenish
wackestone areas. These have a more chaotic fabric
and likely represent burrows, with a bioturbation index
of 4 (sensu Droser & Bottjer 1986). Bioclasts (mainly
trilobites, but also brachiopods and echinoderms) are
thickened and fragmented (Fig. 12A–B).

A well-defined pyrite impregnated lamina is pre-

cent at 0.24 m above the formational base, and less

distinct ones occur at regular intervals up to 1.0 m

above the base of the formation, where three conspicu-

ous anastomosing pyrite laminae are developed (Fig.

11B–C). These coincide with the M. simon/Megistaspis

limbata zonal boundary (Nielsen 1995). Bedding planes in the middle part of the mem-
er are marked by thin, undulating clay seams. In the
uppermost 0.2 m of the unit these are instead de-
veloped as up to centimetre thick marls (Fig. 11D).

The lithology alternates between wackestone and
packstone, comprising a mixed facies as defined by
Harris et al. (2004), indicating a depositional setting in
the middle shelf. This is concordant with the ichnofab-
ric where the bioturbation index, as well as the ubiqui-
tous bored skeletal clasts, suggests a low sedimenta-
tion rate and a high degree of reworking. Conodont
studies by Rasmussen (1991) shows considerably
higher conodont element densities (up to 700 speci-
mens/0.7 kg) in the lowermost 0.5 m, being dominated
by shale. The beds are stacked in a cyclic pattern and
thicknesses of individual carbonate beds range in be-
tween 1 and 20 cm, but the thicker ones are cut by
dissolution seams.

Trilobites and brachiopods represent important
constituents of the fauna, but echinoderms and bryozo-
ans are more important in the Lysaker Member than in
the underlying Hukodden Member.

4.1.4.3 Svartodden Member

The rhythmic beds of the Lysaker Member are con-
trasted by the compact dark grey limestones of the
overlying Svartodden Member (Fig. 13A). Bio-
stratigraphically, the base of the Svartodden Member
belongs to the Lenodus variabilis zone while the top
yields conodonts of the Lenodus crassus zone
(Rasmussen et al. 2013). The unit is homogenous at
first glance, but shows a variation in faunal and litho-
logical composition upon closer inspection; the unit is
relatively fine grained in the lowermost 0.5 m, being
dominated by wackestones. Above this interval, pack-
stone makes up the lithology and endoceratid conches
are common, especially in the mid-upper part.

The basal 0.2 m thick bed constitutes a wackestone
with Thalassinoides burrows. Brögger (1882) noted
characteristic, millimetre sized siltstone phosphorite
grains and dubbed this the ‘Porambonites bed’ due to
its high content of this brachiopod.

Endoceratids appear for the first time at 0.6 m
above the base of the member and are frequent up to
2.15 m from the base (Figs. 10,13B. These conches
frequently show signs of dissolution in their upper
half, and these omission surfaces indicate that the wa-
ter was not saturated with respect to calcium car-
bonate. Siphuncles are often, but not always oriented
way down (Fig. 13C).

Whereas Hansen (2011) interpreted the Svartodden
Member as shallow water deposits, well in the reach of
storm waves but not quite above fair weather wave
base, Nielsen (2004) considered these pure limestones
to be the product of a drowning event that shut off
Fig. 11. Hukodden and Lysaker members. A: Twin hardgrounds near the base of the Hukodden Member. B: Overview of the massive limestones of the Hukodden Member. C: Conspicuous pyrite impregnated laminae in the middle of the Hukodden Member, marking the base of M. limbata Zone (Nielsen 1995). D: The Lysaker Member is characterized by lime-marl alterations, with the more nodular facies to the right being more prone to erosion. Stratigraphic up is to the right. The Volkhov-Kunda boundary bed is visible in the top right corner.
Fig. 12. Photomicrographs from the Huk and Elnes formations. A: Trilobite wackestone with pockets of packstone, Hukodden Member. Packstone shells are unbroken, in contrast to those of the wackestone B: Wacke- and packstone, Hukodden Member. Ubiquitous fine grained pyrite. C: A lime-band of the Lysaker member of trilobite and echinoderm wackestone. D: Close-up of (C). Note shell fragments also in marly, lower left part. E: Well sorted and bioturbated trilobite wacke-packstone, Svartodden Member. F: Trilobite and echinoderm packstone, Svartodden Member. G: Trilobite wackestone, Helskjer Member (Elnes Formation). Note abundance of bored bioclasts.
Fig. 13. The recently re-excavated VEAS section at Djuptrekkodden peninsula, note hammer for scale. The tripartite development of the Huk Formation is very well expressed. B: The compact packstones of the Svartodden Member feature abundant endoceratids with omission surfaces. C: The omission surface of this endoceratid is arranged obliquely to bedding, indicating post-depositional disturbance. D: Hardground with amphora-like borings near the top of the Svartodden Member.
clastic supply. The often well sorted sediments, the coarse grain size and the low degree of lime mud in the middle and upper parts (Fig. 12E) support the former opinion. Indicative of a depositional setting below fair weather wave base is the fact that most, but not all, nautoidal siphuncles are arranged close to bottom position. This suggests that the conches were undisturbed by wave action at least until they had time to become securely fastened by surrounding sediment (Reyment 1968).

A facies shift is seen 0.3 m below the top of the formation. This part does contain a few endoceratids but is characterised by frequent stylolites. These occur at regular, approximately 3 cm, intervals and show amplitudes of 1–3 cm. In this interval, more precisely at 2.35 m above the base of the member, a possible hardground is developed (Fig. 13D). It is not heavily mineralised, but feature frequent amphora-like borings with narrow necks (cf. Trypanites), extending typically 4 cm below the surface.

4.1.5 Helskjer Member (Elines Formation)
The basal metre of the Helskjer Member was exposed at the VEAS road cut, but only very poorly so after the re-excavation during the summer 2014. The lithology makes the unit susceptible to erosion: wavy bands and nodules of limestone interbedded with marls. The bioclast content is dominated by bored trilobite shells, but echinoderms and brachiopods also contribute. The basal part of the formation is composed of wackestones, but bioturbation is visible as channels filled with packstone (Fig. 12F–G). Higher up, they grade into a more fine grained facies, characterised by mud to wackestones with only small and highly fragmented bioclasts.

The strata display a general fining upward trend, with a mixed wacke-packstone facies at the base of the unit and a mud supported facies seen in the upper part. The Helskjer Member is also overlain by the black shales of the Sjostrand Member (Owen et al. 1990, Hansen 2009). Overall, this facies association indicates a transgressive development during the time of deposition.

4.2 Brunflo #2 core
The Brunflo #2 core was recovered by the Swedish Geological Survey in 1970 and preserves strata ranging from the Furongian Alum Shale to the Dariwilian Segerstad Limestone (Fig. 14).

4.2.1 Ceratoppye Limestone
The Ordovician strata of the Brunflo core start with the Ceratoppye Limestone, a thin glauconite-enriched unit which is separated from the underlying Cambrian limestones by a sharp discontinuity surface at -41.20 cm. The topmost Cambrian limestone in the core yields abundant olenid trilobite fragments that can be assigned to the Jiangshanian Stage (Wu et al. in press). Conodont studies performed in the Brunflo area by Sturkell (1991) indicated that Ceratoppye Limestone belong to the upper Paltodus deltifer zone, corresponding to the Tr2 stage slice (Pärnaste et al. 2013). In other words, a substantial hiatus is inferred between the Cambrian and Ordovician strata.

The Ceratoppye Limestone is only 0.05 m thick and was first recognised as a separate formation in the area by Sturkell (1991). The formation comprises two thin glauconitic beds (Fig. 15A), representing wackestone to packstone dominated by vermicular and fractured glauconite grains. The lowermost bed is also rich in trilobite and brachiopod skeletal clasts as well as a few pyrite grains. In contrast, the upper bed is completely devoid of bioclasts and the glaucony takes on a significantly darker green colour. The beds are separated by a distinct but irregular surface. A darker green colour indicates higher maturity of the glauconite in the upper bed (Odin & Matter 1981); this conclusion is supported by the performed SEM-studies (Table 2). Potassium content is high in the glaucony of both beds; the lower bed has an average content of 9.03% K, while the upper bed has a slightly higher average content; 9.40%.

The glauconitic minerals in the Caratoppye Limestone are fractured. This feature indicate (although not unequivocally) an autochtonous origin of the grains, as these otherwise would break along these weakness zones (Amorosi 1997). Nonetheless, a certain degree of reworking is inferred by the findings of abraded conodont elements (Sturkell 1991).

4.2.2 Latorp Limestone
The 2.5 m thick Latorp Formation encompasses dark grey limestones with subordinate grey shales. Biostatigraphically, the formation spans the Paroistodus proteus zone and ranges up into the basal Prioniodus elegans zone (Sturkell 1991).

The carbonate facies comprises extensively recrystallised fine grained nodules, surrounded by an often better preserved wackestone matrix (Fig. 15B). The basal decimetre contains glauconite grains which make up below 5% of the rock, but glauconite is otherwise absent in the formation. Up to six hardgrounds occur at regular intervals between -40.36 and -39.91. These are corrosional in character and have amplitudes of up to 4 cm (Fig. 15C).

4.2.3 Tøyen Shale Formation
The Tøyen Shale Formation is 6.3 m thick in the Brunflo #2 core and consists of dark grey shales with interbeds of trilobite wackestones (Fig. 15D). The base of the Tøyen Shale Formation is diachronous in the Jämtland area, beginning in the Tetragraptus phylograptooides Zone in the Fläsjön area whereas the section in Kloxåsen starts in the Phyllograptus densus Zone (Jaanusson et al. 1982).
Fig. 14. Log of the Brunflo #2 core with carbon isotope curve, partly based on Wu et al. (in press).
Table 2A. Elemental composition of glauconite in the basal bed of the Ceratopyge Limestone.

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Table 2B. Elemental composition of glauconite in the upper bed of the Ceratopyge Limestone.

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The limestone beds of the formation are commonly less than 0.05m thick, but at -35.5m a 0.25 m thick bed can be observed. Above -34 m, the content of bioclasts increases both in the shale and in the limestone facies. This makes the shales turn a lighter shade of grey. Gradually, the shale content also decreases up to the base of the overlying Lanna Limestone.

4.2.4 Lanna Limestone

The 7.2 m thick Lanna Limestone comprises compact trilobite wackestones. It contains a diverse shelly fauna and may biostratigraphically be divided into three biozones of *Megistaspis lata* (equivalent to *Megistaspis polyphemus*), *Megistaspis simon* and *Megistaspis limbata*, in ascending order (Karis 1998; Jaanusson et al. 1982).

The Lanna Limestone takes on a red colour but ubiquitous greenish *Thalassinoides* burrows contrasts strongly against the host rock. The formation is partly nodular owing to the presence of clay seams and marl beds.

Hardgrounds occur at two levels in the formation. The first level is a 7 cm thick interval with repeated hardgrounds (Fig. 15E) which can be found at ~30.6 m. This interval may possibly be an expression of the wide spread hardground complex 'Blommiga bladet' (Lindström 1979). Another hardground is present at -28.58 cm. It is manifested as a sharp facies shift from limestone to shale, with up to 5 cm deep burrows or borings protruding down from the upper surface (Fig. 15F). This hardground may correlate with ‘Blodläget’ (Lindström 1979).

4.2.5 Holen Limestone

The Holen Limestone can be divided into two informal members on the basis of its expression in the Brunflo #2 core; the lower one of these is characterised by rhythmic lime-marl alternations whereas the upper one is a compact, thick bedded limestone. Löfgren (1978) defined the lower one as the Flåsjö Limestone and the upper one as Järvsand Limestone. The lower part of the formation preserves a rich fauna belonging to the *Asaphus expansus* Biozone (Jaanusson et al. 1982).

The base of the Holen Limestone is marked by a limestone clast conglomerate over a weakly defined erosional surface in many parts of Jämtland (Karis 1998). Such an interval is not evident in the Brunflo core, but at ~23.7 m, a 0.05 m thick glauconitic grainstone bed with a sharp, scoured base can be observed (Fig. 15G); a level which likely correlates with the conglomerate, and which would thus mark the base of the Holen Limestone. The glaucony in this bed is represented by light green, elongate grains. The base of the Holen limestone is also marked by the transition from *M. limbata* to *A. expansus* biozones, i.e. the Volkov-Kunda transition in the Baltoscandian regional stages. This interval marks a pronounced sea level fall throughout Baltoscandia (Lindskog et al. 2014). This is concordant with the scoured base of the glauconitic bed discussed above, which indicates a significant sea level change.

The lower informal member, the rhythms of the Flåsjö Limestone, can be subdivided into a lower part with medium thick limestone beds and an upper part with lenticular limestone nodules. The lower part comprises medium bedded, dark grey, wavy continuous bands of brachiopod and trilobite wackestone (Fig. 15H), interbedded with marly shales. This facies is overlain by discontinuous nodular limestones at ~16.5 m, which show thicker marl interbeds. The marl beds gain a progressively darker colour higher up, peaking at ~11.80 m. After this they become more shell rich and less dark.

The upper informal member, the Järvsand Limestone, is instead characterised by compact and thick dark grey limestone beds. Shale content is low. Just as in the Flåsjö limestone, the fauna is dominated by trilobites and brachiopods, but an interval at -7.80 to -6.40 m yield abundant endoceratid conches.

4.2.6 Segerstad Limestone

The Segerstad Limestone is not defined on lithological grounds; it is a topostratigraphical unit in which the base is defined by the appearance of the trilobite *Asaphus platyurus* (Jaanusson 1960). As such, it is hard to pinpoint the base of the Segerstad Limestone since no trilobite biostratigraphical work has been performed on the Brunflo core. Nonetheless, Larsson (1973) mentions several hardgrounds just above the base of the Segerstad Limestone. Three haematite impregnated hardgrounds of corrosional type occur at -4.90 m in the Brunflo core (Fig. 15I), and these may well correlate to
Fig. 15. Polished and scanned slabs from Brunflo #2 core, from Wu et al. (in press). Field of view is 3 cm wide. (A) The two beds of the glauconitic Ceratopyge Limestone. The glaucony of the upper bed is darker and have higher potassium content. Two pseudosparitic nodules in the upper part represent the lowermost Latorp Formation. (B) Recrystallized limestone clasts set in a dark grey wackestone matrix. (C) The middle part of the Latorp Limestone is characterized by corrosional hardgrounds, note the marked and irregular morphology of these two. (D) Typical lithology of the limestone beds of the Tøyen Formation. Possible escape trace can be seen in the upper left corner. (E) Hardground complex tentatively correlated with ‘Blommiga bladet’ (sensu Lindström 1979) in the lower part of the image. (F) Well defined hardground possibly corresponding to ‘Blodläget’ (Lindström 1979). (G) Glauconitic bed with sharp scoured base, interpreted as the Volkov-Kunda boundary. (H) The lowermost part of the Holen Formation comprised thick bedded wackestone interbedded with marls. (I) Three haematite impregnated hardgrounds approximate the base of the topostratigraphical Segerstad Limestone in the core.
those described by Larsson (1973). They can thus serve to approximate the base of this unit. Lithologically, the Segerstad Limestone is, just as the upper part of the Holen Limestone, dominated by thick bedded trilobite wackestones. The rock is generally red, except where bioturbation has altered the rock green or grey. Minute pyrite crystals can be found dispersed throughout the unit.

5 δ13C-record

While the global record of the Upper Ordovician feature several well-studied high amplitude carbon isotope excursions, the Lower and Middle Ordovician is characterised by having smaller isotope shifts and a more subdued generalised curve (Bergström et al. 2009). The most prominent excursion in the Lower and Middle Ordovician is the mid-Darriwilian carbon isotope excursion (MIDICE; e.g. Meidla et al. 2004; Ainsaar et al. 2010; Schmitz et al. 2010), but lately, several smaller scale excursions have been described from Baltoscandia (Lehnert et al. 2014). Four of these are of interest of the present study, namely LTNICE (Late Tremadocian Negative Isotopic Carbon Excursion), BFICE (Basal Floian Isotopic Carbon Excursion, which marks the onset of the Floian-Darriwilian rise), BDNICE (Basal Dalipingian Negative Isotopic Carbon Excursion) as well as the LDNICE (Lower Darriwilian Negative Isotopic Carbon Excursion). In Lehnert et al. (2014), the LDNICE is present in the lower Darriwilian (lower Kunda) and precedes the MDICE, which shows peak values in upper Dw2 and basal Dw3 (Ainsaar et al. 2010).

5.1 Slemmestad

The δ13C_carb data from Slemmestad (Fig. 10, Tables 3–4) show strongly fluctuating and surprisingly low δ13C_carb-values, and should thus be treated with caution. No clear trend can be seen in the dataset. When compared to the global δ13C_carb curve by Bergström et al. (2009; which for this interval is based on data from Baltica by Kaljo et al. (2007)) the results presented here are about 2 ‰ lower, i.e. significantly shifted to the negative side. Similarly low and fluctuating results have been presented for the Svtavdottir Member from another locality in the Oslo area (Guttormsen 2012). The poor data likely reflects diagenetic disturbances of another locality in the Oslo area (Guttormsen 2012). Samples from Slemmestad were extracted from pristine rock that likely has not been altered by humic fluids. Alteration of the isotopic signature may in this case be derived from tectonic overburden during the subsequent Caledonian Orogeny. Conodont colour alteration palaeotemperometry shows that temperatures reached 300 °C during the orogeny (Bergström 1980).

The δ13C_carb-curve from the Tøyv Shale Formation is more stable and is thus inferred to be more reliable. The data shows a generally decreasing trend throughout the Hagastrand Member, with some deviations (Fig. 9, Table 5). The first appearance of Tetragraptus phyllograptoides marks the base of the Floian, and this interval is slightly below the base of the Galgeberg member (sensu Owen et al. 1990). This coincides with a 3.5 ‰ positive excursion, after which a shift to increasing values is seen. An excursion peak followed by a rising trend fits with BFICE and the onset of the Floian-Darriwilian rise of Lehnert et al. (2014). Caution should be exercised however, as this would-be excursion is represented by a single sample. A higher resolution would be preferable to confirm this excursion. Furthermore, δ13C_carb-values do not always correspond to similar trends in δ13C_org as they are controlled by different factors (cf. Young et al. 2008).

The upper exposure shows a slightly increasing trend but no clear excursions. The uppermost sample show very high values but this sample was very rich in carbonate, and it is plausible that carbonate leaching was not sufficient during treatment in the lab of this particular marl sample.

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Table 3. δ13C_carb data from the Bjørkåsholmen Formation.
Table 4. $\delta^{13}C_{\text{carb}}$ data from the Huk Formation

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<th>Height above base of $\delta^{13}C_{\text{carb}}$ Sample member (V-PDB)</th>
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Table 5A. $\delta^{13}C_{\text{carb}}$ data from the lower exposure, Tøyen Shale Formation

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Table 5B. $\delta^{13}C_{\text{carb}}$ data from the upper exposure, Tøyen Shale Formation

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</tr>
<tr>
<td>Slemmestad</td>
<td>B42</td>
<td>-0.34 -28.97</td>
<td>Galgeberg</td>
</tr>
<tr>
<td>Slemmestad</td>
<td>B43</td>
<td>-0.02 -25.44</td>
<td>Galgeberg</td>
</tr>
</tbody>
</table>
5.2 Brunflo

As opposed to the carbon isotope studies from Slemmestad, the Brunflo data shows reliable values throughout the studied interval. The Brunflo core #2 shows a general increasing trend in $\delta^{13}C_{\text{carb}}$ starting with the values of approximately -0.5‰ of the Latorp Formation, and ending at over +1.3‰ in the Segerstad Limestone, with a number of smaller fluctuations and subtrends in between (Fig. 14). The Latorp Formation records a dip which likely corresponds to the LTNICE of Lehnert et al. (2014). The curve rises and shifts to positive values at the transition to the overlying Tøyen Shale Formation, continues to rise and culminates in a peak near the top of the formation, reflecting the lower part of the Tøyen-Darriwilian rise (Lehnert et al. 2014). Values drop to below zero between the Tøyen and Lanna formations, possibly representing the Basal Dapingian excursion of Lehnert et al. (2014), an interesting interval as it is just below the regionally important hardground complex ‘Blomniga bladet’. The isotope data from the Lanna Limestone show a bulge, starting at low values, with higher values in the middle that drop again near in the transition to the Holen Formation. This interval marks the transition between the Volkhoov and Kunda regional stages. Meidla et al. (2004) also documented a negative excursion at this interval in the Gullhögen quarry in Västergötland, just a metre below the conspicuous ‘Täljsten’ interval.

Values in the lower member of the Holen Formation vary between 0 and 0.5‰, showing a series of fast shifts between -19 and -17 m that may have correlational importance, as the ‘Täljsten’ interval in Meidla et al. (2004) is also marked by a series of fast shifts. A fast shift exceeding 0.6‰ can also be seen in the data from Tingskullen of Wu et al. (submitted) at the transition from Lenodus variabilis and the Yangtzeplacognathus crassus zones, which is time equivalent to the ‘Täljsten’ (Mellgren & Eriksson 2009).

A distinct negative excursion with a magnitude of 0.5‰ is present at the transition to the upper member of the Holen Limestone where it marks the onset of the MDICE. The excursion has previously been interpreted as the LDNICE (Wu et al. in press) but the level is likely too high up in the stratigraphy as the LDNICE was originally defined by Lehnert et al. (2014) to be in the basal Kunda and the ‘Täljsten’ interval. The LDNICE as originally defined by Lehnert et al. (2014) is likely best represented by the ‘Täljsten’ shift mentioned above.

The upper member of the Holen and the Segerstad limestones show a strong and steady increase in $\delta^{13}C_{\text{carb}}$-values, going from 0 to 1.3‰. The stratigraphic position as well as the magnitude of the excursion corresponds to the rising limb of MDICE (Meidla et al. 2004; Ainsaar et al. 2010). This excursion is characterised by three subpeaks, and the first of these is represented in the uppermost part of the Holen Formation. The second and highest peak of MDICE is not preserved in the core interval.

6 Comments to lithology

6.1 Pseudomorphs, recrystallisation and possible implications

Major recrystallisation characterises the Bjørkåsholmen Formation and the Latorp Limestone in the Brunflo core section and presumed pseudomorphs of evaporitic minerals are also found in the Bjørkåsholmen Formation and the Hagastrand Member in Oslo. The most conspicuous of the presumed evaporite pseudomorphs are those found in ‘the main limestone bed’ of the Bjørkåsholmen Formation in Oslo. Here, the spatial relationship between the presumed evaporite pseudomorphs and limestone nodules indicates that the evaporite crystals grew within the sediments, thereby displacing the nodules (esp. in Fig. 8B). Demicco & Hardie (1994) writes that sediment displacive growth of evaporites unequivocally means that evaporitic conditions has occurred, but it is uncertain whether these were met closely after deposition or later during diagenesis by saturated deep brines. According to Kendall & Harwood (1996) supersaturated brines are only likely to be formed at the brine-air interface. This would mean that the crystals formed penecontemporaneously to deposition and that depositional depth cannot be more than a few metres deep. This is because deep bodies of brine almost always are stratified, and concentrated surface brines do not descend to the basin floor. Furthermore, in modern environments, sulphates form in marginal marine settings and become more dominated by halite shorewards (Kendall & Harwood 1996). We do not see signs of halite – no cubic crystals – indicating that the Bjørkåsholmen Formation was likely deposited in a shallow environment.

The crystal mushes indicate that gypsum was abundant. This type of displacive crystal growth normally forms in association with algal mats or fine grained sediment (Kendall & Harwood 1996). The latter is more likely as neither filaments, fenestrae nor other indications (Tucker 2001) of algal mats are found. These crystals are set in a matrix which shows a transition of pseudospar to microspar and micrite. This phenomenon is briefly discussed in Scholle & Ulmer-Scholle (2003). They write that the mechanism behind is not fully understood, but that such recrystallisation is often associated with early meteoric exposure and tectonic stresses or large scale, syndepositional, glacioeustatic sea level fluctuations.

While Egenhoff et al. (2010) mentioned heavy recrystallisation of the Norwegian successions, they did not discuss why this is the case. According to them, crystal sizes reflect original grain size, i.e. the fine grained recrystallised rock derive from mud- or wackestones whereas the coarse grained ones reflect packstones. By this way of reasoning, they recognise fourteen deepening upwards cycles in the strata, based on the perceived fining upwards trend in the carbonates. In my opinion however, recrystallisation of the for-
formation is so severe that such conclusions cannot be drawn. This is evident from both texture and low δ18O-values (cf. Wu & Wu 1996). Crystal sizes of recrystallised material depend not only on primary clast size but also on mineralogy; abundant unstable primary aragonite allows for a high degree of secondary precipitates (cf. Lasemi & Sandberg 1984; Munnecke 1997).

The Latorp Limestone also shows recrystallisation structures similar to the ones seen in the Bjørkåsholmen Formation, and may indicate that evaporitic and restricted conditions were more widespread in the Early Ordovician of Baltoscandia than previously thought. If so, why is this only at this level and not higher up? After all, climate must have been warmer as the continent was approaching the equator (Cocks & Torsvik 2005). According to Nielsen & Schovsbo (2013) the basin was silled, and a sea level drop could potentially produce lagoonal evaporitic conditions. Furthermore, the Early Ordovician represents a very warm period and is characterised by greenhouse conditions. However, sea surface temperature gradually cooled and reached modern equatorial sea surface temperatures during the Middle Ordovician, reflected by a drop of ≈10°C from the middle Tremadocian to the Darrxiwalian (Trotter et al. 2008). Global climate change with major cooling of the Earth’s oceans may thus be the driving factor behind the absence of evaporitic facies higher up in the Baltoscandian Ordovician successions.

6.2 Glaucony

Glauconite is, in the strict sense, a green, potassium and iron rich ferric micaceous mineral. It forms in marine sediments at intervals of slow deposition and often occurs in a granular habit (e.g. Hugget 2005). As Odin & Matter (1981) noted however, the term ‘glauconite’ is commonly both used for green grains in general as well as the facies in which it occurs. To avoid confusion, they proposed that the term ‘glauconitic minerals’ be used for the clay minerals that characterise the glaucony facies, i.e. a clay mineral family with a pronounced variation in potassium content, the end-members of which being a K-poor glauconitic smectite and a K-rich glauconitic mica. They also proposed the term ‘glaucony’ be used for the blue-green marine facies characterised by these green glauconitic minerals (see also Odin 1988).

Small pores of various substrates, ranging from skeletal debris to lithic grains to faecal pellets, provide favourable microenvironments for glauconisation. These semi-confined conditions provide a chemistry that differs from both that of the sea water and the sediment pore-water as ion interchange with the surroundings does occur, but at a restricted rate (Odin & Matter 1981; Odin & Fullagar 1988). As iron is not mobile during oxidising conditions, and as iron tends to form pyrite in truly reducing milieus, glaucony is inferred to form in sub-oxic, slightly reducing environments during shallow burial (Kelly & Webb 1999). Sea water contains only traces of Al, Si and Fe and the chemical constituents of the glauconitic minerals may be supplied from the substrate grain as it dissolves (Clausen et al. 1992).

According to Odin & Matter (1981), the first mineral to form is a K-poor glauconitic smectite and as time progresses more and more smectite is formed. At the same time, the crystals that have already formed undergo a maturation process in which they gradually incorporate more K and become less expandable. In this maturation process the glauconitic smectite recrystallises into glauconitic mica and the original substrate grain tends to dissolve completely. K2O-content of glaucony in the nascent stage is between 2–4 wt % but rises and exceeds 8 wt % in the highly evolved stage (Odin & Fullagar 1988).

The mineralogy of glauconitic minerals is preferably studied using X-ray diffraction (XRD), as this kind of study allows for distinction between mica and smectite (Odin & Matter 1981). On the other hand, most of the physical properties, including e.g. XRD-patterns, density and paramagnetism of glaucony, can be correlated with the potassium-content (Odin & Fullagar 1988). An XRD-device has not been available to the present study, but energy-dispersive X-ray spectroscopy (EDX) was employed instead. While this method cannot differentiate between clay mineral lattices, it allows measurement of potassium content.

The potassium content of the glauconitic minerals of Bjørkåsholmen formation is on average 7.43% and can thus be considered to be of evolved type, a product of a maturisation process which may take over 105 years (Odin & Fullagar 1988). On the other hand, these minerals have likely been diagenetically altered (Bjørlykke 1974) and the results should thus be treated with caution. The data from the Ceratopyge Limestone are more reliable and show that the upper bed is more evolved than the lower bed, indicating a longer time of slow-deposition when this bed formed. Both beds can be considered as highly evolved, indicating a period of slow-deposition approaching 106 years for each of these beds (Odin & Fullagar 1988).

6.3 Limestone-marl alternations

The Lysaker Formation is best described as a limestone-marl alternation sensu Munnecke & Samtleben (1996). Here, the term ‘marl’ is not used in a strict sense, i.e. as a specific lime/clay ratio, but as a designation for the part of the rock that weathers out more easily. Limestone-marl alternations are thought to form in the shallow, early diagenetic realm due to dissolution of aragonite and subsequent precipitation as calcite. At a certain depth (‘Aragonite Solution Zone’ or ASZ), the pore water chemistry renders aragonite unstable and it is dissolved. This zone where dissolution occurs will come to form the marly horizons. The calcium carbonate migrates upwards where it reprecipitates as calcite, leading to what will become limestone layers (Munnecke & Samtleben 1996). On the other hand, aragonite dissolution may also occur at the sediment-water interface (Palmer & Wilson 2004). In this case,
aragonite is dissolved but does not reprecipitate as cement; instead the ions go into solution in the sea water. Nonetheless, in the model of Munnecke & Samtleben (1996), the lime-marl alternations are not formed due to inhomogenieties in the precursor sediment but because to the presence of unstable aragonite. These alternations can thus not be used to infer climatic changes due to for instance Milankovitch fluctuations (Westphal 2006).

Preliminary palynological studies of various rhymites of the Oslo Region, including the Lysaker Member, have been presented by Amberg et al. (2013). These show that the palynomorph assemblages of limestones and mudstones do not differ significantly. Furthermore, findings by Egger et al. (2013), who studied the Skogerholmen Formation of the Upper Ordovician in the Oslo Region, show that ratios of diagenetically insoluble elements (in this case Ti and Al) are essentially the same in both lithologies. This would not have been the case if the rhymites were a primary feature and with significant differences in clay mineralogy composition at the time of deposition of lime and marl layers. Together, these two studies lend strong support to the hypothesis that the limestone-marl alternations derive from diagenetical processes and that these sediments were originally more or less homogenous.

The ratio of the marl and limestone thicknesses of marls and limestones depends on which calcium carbonate polymorph dominated the precursor sediment; a high content of aragonite would give rise to thicker limestone beds (Munnecke et al. 2001). The Lysaker Member does not only show rhythmic beds, it also shows cyclicity at a larger scale, with limestone beds being thicker and more prominent at certain levels. If the model of Munnecke & Samtleben (1996) is correct, this suggests that more aragonite as compared to calcite was precipitated at the time these levels became deposited. Which polymorph to be precipitated depends both on the marine Mg:Ca ratio as well as the temperature of ambient waters, with aragonite formation being favoured at high Mg:Ca ratios and high temperatures (Balthasar & Cusack 2015). The larger scale cyclicity that can be observed in the Lysaker Member does not only show rhythmic beds, it also shows cyclicity at a larger scale, with limestone beds being thicker and more prominent at certain levels. If the model of Munnecke & Samtleben (1996) is correct, this suggests that more aragonite as compared to calcite was precipitated at the time these levels became deposited. Which polymorph to be precipitated depends both on the marine Mg:Ca ratio as well as the temperature of ambient waters, with aragonite formation being favoured at high Mg:Ca ratios and high temperatures (Balthasar & Cusack 2015). The larger scale cyclicity that can be observed in the Lysaker Member may thus be a function of climate fluctuations.

7 Chemostratigraphy and environmental implications

Variations in $\delta^{13}C$ of the world’s oceans through time have foremost been used for correlation. However, the $\delta^{13}C$ is mainly dependent on the distribution between organic carbon and carbonate rocks, and is thus directly linked to the global carbon cycle and the geosphere development (Saltzman & Thomas 2012).

Numerous factors influence the $\delta^{13}C$-development however, and the connections are not always clear-cut. Volcanism, the proportion of carbonate relative to silicate weathering and the composition of terrestrial vegetation are a few of the factors that influence the carbon isotope development (Kump & Arthur 1999). Numerous episodes of extinctions or biotic turnover have also been linked to carbon isotope excursions (Saltzman & Thomas 2012).

The Ordovician Period records an unparalleled diversification among marine invertebrates. This evolutionary event has been coined ‘The Great Ordovician Biodiversification’ or GOBE (Webby et al. 2004), and its importance is on a par with the ‘Cambrian explosion’. Whereas the latter generated a wide disparity and larger number of taxonomically higher ranking taxa, it was the former that provides the sheer biomass and any substantial biodiversity at genus and species level (Harper 2006). Tiering complexity increased and a number of new niches were also occupied, which for instance can be seen in the coeval increase of intensity and diversity of carbonate substrate bioerosion (‘Ordovician Bioerosion Revolution’ of Wilson & Palmer 2001, 2006).

Increased tiering and the occupation of new niches associated with the onset of GOBE likely led to a shear increase in total biomass and, with that, increased burial of organic material. Organic material is enriched in light $^{13}C$ as compared to $^{12}C$. Large scale burial of organic matter thus leads to that the oceans become depleted in the lighter isotope, giving a long term increasing trend in the $\delta^{13}C$-record (Knauth & Kennedy 2009). It has also been suggested that the increase in carbon burial rate lead to climatic cooling, which further stimulated the development of the GOBE (Trotter et al. 2008, Zhang et al. 2010). The Floian marks the start of a long-term rise in the carbon isotope record, the Floian-Darriwilian rise (Lehnert et al. 2014), a development which eventually culminates in the MDICE. Does this timing fit with the timing of the GOBE?

To begin with, the biodiversification event was by no means a continuous steady increase in species and genera, but followed a stepwise pattern, varying for different taxonomical groups (Webby et al. 2004; Harper 2006). To some extent, these biodiversity peaks reflect sea level; graptolites peak at high sea levels whereas the opposite is seen with ostracodes and brachiopods. This gives a general trend where high sea levels correspond with low biodiversity and vice versa (Hammer 2003; but this may well just be a taphonomical effect, as the shallow water sediments deposited during highstand are preferentially eroded away, and thus also the fossils). On Baltica, the GOBE is expressed as four distinct peaks in diversity: one at the beginning of the Floian Stage, one in the upper Dapingian/lower Darriwilian (Langevoja regional sub-stage), one in the late Sandbian and finally one in the Katian. Conversely, the Dw2 and Dw3 record a prominent dip in diversity (Hammer 2003).

The onset of the Floian-Darriwilian rise thus matches with the onset of the GOBE, whereas the MDICE corresponds to a diversity dip. Interpreting environmental signals from the $\delta^{13}C$-record is by no
Fig. 16. Correlation of Lower–Middle Ordovician strata across western Baltoscandia, based on carbon isotope stratigraphy, widespread hardground complexes and biostratigraphy. See text for detailed discussion.
means straightforward, and a simple correlation between diversity and carbon isotope trends does not exist.

8 Correlation and depositional history

To correlate the sections in Oslo and Jämtland, an integrated approach is adopted here by combining biostratigraphic data with carbon isotope trends and microfacies analysis, with the main results presented in Figure 16. The sections in Slemmestad and Brunflo both represent relatively deep marine facies on the western margin of Baltica, although Slemmestad had a generally deeper setting with a higher siliciclastic influx. The Norwegian successions are considerably more expanded, especially during Tremadocian and Floian times.

As no analogues of the vast, epicratonic seas that characterised the Early Palaeozoic exist today, sea level interpretation and application of sequence stratigraphy for these successions is problematical. Nonetheless, various techniques may be used to deduce palaeo-sea levels, including sedimentological evidence, physical proxies (such as regional relief), ecological and biological assemblages as well as geochemical data of oxygen isotopes (Munnecke et al. 2010).

Studying the Upper Ordovician successions of western Estonia, Harris et al. (2004) used the depositional texture in the classification scheme of Dunham (1962) as a basis for a model to interpret epicratonic seas. In their model, facies belts shift laterally from shelf to basin with grain supported facies being deposited in the shallow shelf, mixed facies in the middle shelf, mud-supported facies in the deep shelf and slope and finally black shale facies in the basin (Fig. 17). While this makes sense intuitively, one must bear in mind that it is a simplified model and that carbonate sediments are biologically formed sediments – i.e. they are affected not only by sea level but also for instance by climatic factors. Nonetheless, the simplicity of the model makes it useful for interpretations.

Ordovician sea level curves have been reconstructed for several continents, e.g. Laurentia (Ross & Ross 1992, 1995), Sibiria (Kanygin et al. 2010) and the Yangtze Platform (Su 2007). Long ranging sea level curves for the Ordovician of Baltoscandia include those by Nielsen (2004) and Dronov et al. (2011). Nielsen (2004) focussed on the Scandinavian, distal parts of the basin whereas Dronov et al. (2011) based their curve on the proximal parts in the shallow-water settings of Estonia.

These roughly correspond to one another in the Tremadocian and Floian but differ considerably in the Dapingian and Darriwilian; Nielsen (2004) inferred a lowstand at the same time as Dronov (2011) perceived a highstand. This contradiction is partly based on the fact that their curves are based on the distal and proximal parts of the platform respectively but also because they used different approaches to reconstruct sea level. Dronov et al. (2011) base their assessments on two assumptions that major regional unconformities represent forced regressions and that the expansion over a wider area of deep water facies, such as marine red beds, reflect transgressions. However, it may be argued that, as regional unconformities expressed as hardgrounds represent highly condensed strata (Flügel 2010), sea level falls cannot be deduced from these.

8.1 Tremadocian 2

The interval of interest for the present study starts in the Tremadocian 2 time slice with the deposition of the Bjørkåsholmen Formation and Ceratopyge Limestone. Both belong to the P. deltifer Zone (Sturkell 1991; Erdtmann & Paalits 1994). This zone has not been documented from Tingskullen, and the time interval represented by the Bjørkåsholmen Formation and the Ceratopyge Limestone may simply be gone due to erosion in association with the karstic level at ~46 m in the Tingskullen core.

The sudden transition from the black, anoxic Alum Shale to the possibly evaporitic limestones of the Bjørkåsholmen Formation suggests a significant sea level drop (Fjelldal 1966, Erdtmann & Paalits 1994, Nielsen 2004). The formation has been interpreted as representing a lowstand systems tract (Dronov & Holmer 1999) or as a falling stage systems tract with a

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Fig. 17. Facies model originally constructed for the Upper Ordovician of Western Estonia (Harris et al. 2004), but a similar large scale morphology and sedimentation pattern likely prevailed in western Baltoscandia during the Lower and Middle Ordovician times. Key to abbreviations: SL = sea level, FWWB = fair-weather wave-base and SWB = storm wave-base.
superimposed lowstand systems tract (Egenhoff et al. 2010). The Björkåsholmen Formation is 1.2 m thick whereas the Ceratopyge Limestone is only 0.05 m in thickness. While the Björkåsholmen rests conformably atop the Alum Shale, a hiatus that spans part of the Cambrian Jiangshanian stage to the basal Tremadocian underlies the Ceratopyge Limestone in Brunflo. This suggests a sea level fall that exposed the Cambrian sediments to erosion in Brunflo but that the magnitude of the fall was not sufficient to expose the Slemmestad strata.

Egenhoff et al. (2010) proposed that the regression was an isostatic event due to uplift in association with an early stage of the Caledonian Orogeny. Erdtmann & Paalits (1994) on the other hand regarded it as eustatic, the Ceratopyge Regressive Event (CRE). It coincides with the transition between the Sauk III and Tippecanoe I megacycles in North America, and Nicoll et al. (1992) considered it identical with the Kelly Creek Eustatic Event recognised in the Georgina and Amadeus basins of Australia. Erdtmann & Paalits (1994) discussed a possible glacial influence, but only controversially discussed tillites have been found from this time. Furthermore, this period was characterised by greenhouse conditions (Trotter et al. 2008) rendering glacial episodes unlikely. What is clear is that the regression coincides with a major faunal turnovers and extinctions in for instance conodonts, graptolites and trilobites in many parts of the world (Erdtmann & Paalits 1994, Albanesi & Bergström 2004).

The CRE was followed by a long term transgression (Nielsen 2004). The glaucony of the Ceratopyge Limestone and Björkåsholmen Formation may thus have formed during the transgression following the sea level fall. A model outlined by Amorosi (1995) predicts that the abundance as well as the maturity of glaucony increases through the upper transgressive systems tract, reaching a peak at the condensed section, and is followed by a gradual decrease in the highstand systems tract. This model fits with the observation of increasing maturity of glaucony in the Ceratopyge Limestone as well as the findings of a few scattered glauconite grains in the basal parts of the overlying Latorp Limestone. The initial precipitation of glauconitic smectite is estimated to occur in a time-span of 1–10 kyr whereas highly evolved glauconitic mica takes 100–1000 kyr to form (Odin & Matter 1981). During this interval, the evolving glaucony must be kept at slightly reducing conditions and not be buried too deep. The presence of highly evolved glaucony thus speaks for a significant break in marine sedimentation (Odin & Matter 1981; Amorosi 1995, 1997; Kelly & Webb 1999). These conditions are more likely to occur during transgressions, when the sedimentation zone moves landwards (Odin & Matter 1981).

8.2 Tremadocian 3 and Floian
The biostratigraphic control on the Hagastrand Member (basal Tøyen Shale Formation in the Oslo Region) is poor, but it is deemed to be coeval with strata belonging to the Megistaspis (Paramegistaspis) planilimbata Zone (Hoel 1999a). This correlates with the upper part of the Latorp Limestone where this Biozone had been documented (Karis & Larsson 1982).

Sequence stratigraphic models are admittedly difficult to apply to deep water facies. Nonetheless, Egenhoff & Maletz (2007) introduced a way to work with sequence stratigraphy also in monotonous shale successions based on taxonomic variations among the graptolites. They found that endemic forms dominate most of the sedimentary interval, but that pandemic deep water taxa form mass-occurrences at certain intervals. They hypothesised that these intervals represented maximum flooding surfaces. And indeed, Erdtmann (1965) also mentions mass occurrences of for instance Clonograptus and multiramous dichograptids at several levels in the Tøyen outcrop in Oslo, but only one of these were specified. This level is dominated by Clonograptus galgebergi and Clonograptus norvegicus and occurs at 3.55-3.68 in the Tøyen outcrop. It is equal to ‘the good bed of the H. copiosus fauna’ (Lindholm 1991) and may correlate to a black shale bed at 7.1 m above the base at Slemmestad, thus likely representing a maximum flooding surface.

The base of the Tøyen Shale Formation in Jämtland is diachronous, and ranges between Tetragraptus phylograptoides and Phylograptus densus biozones (Jaanusson et al. 1982), the first of which is found in the uppermost part of the Hagastrand Member in Oslo and the latter which is found in the middle part of the Galgeberg Member (Erdtmann 1965). The 2.5 mm thick Latorp Limestone thus correlates with at least the basal 11 m of the Hagastrand Member, making the Latorp Limestone extremely condensed. Recrystallisation and pseudomorphs are evident both in Slemmestad and in Brunflo in this interval.

The base of the Floian is marked by carbon isotope excursions in the Slemmestad and Brunflo sections; these are inferred to correspond to the BFICE of Lehnert et al. (2014). The facies transitions from grey shale with silt interbeds to black, anoxic shales in Oslo and the corresponding transition from compact limestone to lime-marl interbeds suggest a continued transgression at this time.

The Floian strata in Oslo is more expanded in Oslo than in Jämtland; the Tøyen Shale Formation is approximately 10 m thick in Jämtland, while the coeval Galgeberg and Slemmestad members is estimated to comprise more than 15 m of strata.

8.3 Dapingian
‘Blommgia bladet’ (‘Flowery sheet’) and ‘Blodläget’ (‘Bloody layer’) are extensive hardground complexes that can be followed over large parts of Scandinavia, from Öland in the SE to Dalarna in the NW, to the Baltic area in the east, serving as important marker horizons across the continent (Lindström 1979). The lowermost of these, ‘Blommgia bladet’, is
characterised partly by abundant amphora-like borings but foremost by its strong colours, ranging from yellow and red to green (Lindström 1979; Ekdale & Bromley 2001). Biostratigraphically, ‘Blommiga bladet’ formed at the base of Baltoniodus triangularis zone, which is the index fossil that marks the base of the Dapingian Stage (Bergström & Löfgren 2008). The exact location of this base has not been pinpointed in the Oslo region as the deep water facies of the Tøyen Shale Formation yield no conodonts (Bergström & Löfgren 2008), and the hardground complex is not developed in the same way in these shaly facies. Nonetheless, the location can be approximated using other biostratigraphic data. The transition from Iso-graptus lunatus to Iso-graptus victoriae biozones is at, or close to, the base of the B. triangularis zone (Bergström et al. 2009; Pärnaste et al. 2013). While I. victoriae has not been documented from the Tøyen Shale Formation, I. lunatus has been reported from a level approximately 3 m below the top of the formation (Erdtmann 1965). The base of the succeeding Huk Formation is within the Baltoniodus navis zone, i.e. its deposition took place well within the Dapingian Stage. The base of the Dapingian and the level of ‘Blommiga bladet’ must thus be within the uppermost 3 m of the Tøyen Shale Formation. This interval contains a multicoloured horizon at 1.3 m below the base of the Huk Formation interpreted as a maximum flooding surface. This level probably corresponds to ‘Blommiga bladet’, but developed in a deeper facies. It is overlain by the grey shales of the informal Slemmestad Member.

‘Blommiga bladet’ is not unequivocally expressed in the Brunflo core either, but it is tentatively recognised as the hardground complex found at -30.6 m, not even a metre above the base of the Lanna Limestone. The basis for correlation of these two beds is thus not very strong. The shift in facies from black shales to grey in Oslo and the reappearance of compact limestone (Lanna Limestone) in Brunflo suggest that the transgression that started in the late Tremadocian had culminated and turned into a regression. ‘Blommiga bladet’ is clearly expressed at -41.95 m in the Tingsskullen core of Öland.

The other hardground level of interest is ‘Blodläget’. This has been identified with a high degree of confidence at 0.17 m above the base of the Huk Formation in Slemmestad, coinciding with the base of Megistaspis simon Zone (Nielsen 1995). A pronounced hardground at -28.58 m in the middle Lanna Limestone may correspond to this level, as M. simon is found in the middle part of this formation (Karls 1998). While the Lower Ordovician strata are considerably more condensed in the Brunflo area than in Slemmestad, the difference is much less pronounced in the Middle Ordovician. The interval between ‘Blommiga bladet’ and the presumed position of ‘Blodläget’ is even more expanded in Brunflo than in Slemmestad. This possibly indicates that the identification of ‘Blodläget’ in Brunflo is incorrect. The confidence in the identification of ‘Blodläget’ in the Tingsskullen core is much stronger. It is expressed as a hematite impregnated surface and the biostratigraphical control, placing it at the base of B. navis is very strong (Wu et al. submitted).

8.4 Darriwilian

The Holen Limestone starts with a glauconitic grainstone bed with an erosional base. This level correlates with the thick packstone bed developed in the middle part of the rhyolites of the Lysaker Formation that marks the Volkhov-Kunda transition (Nielsen 1995), a level that has been associated with a brief sea level fall in Västergötland (Lindskog et al. 2014). This regression should be expressed also in the Tingsskullen core, but detailed sedimentological data of this interval is lacking. At the outcrop at Byrum, northern Öland however, the Volkhov-Kunda boundary is likely represented by goethitic/limonitic ooids in middle of Formation A+B (cf. Stouge 2004). The Volkhov-Kunda boundary marks the start of a ~1.75 Myr long period of increased influx of meteorites which has been linked to a major breakup of a L-chondrite asteroid parent body at ~470 Ma (Schmitz et al. 2001). Strata of the Lenodus variabilis, Yangtzeplacognathus crassus and Microzarkodina hagetiana conodont zones are all enriched in chondrite grains (Schmitz et al. 2008).

The ‘Täljsten’ is a faunally and lithologically anomalous facies developed in the Kunda strata (approximately Dw1-2 boundary) of Västergötland, Sweden, where it is recognisable as an approximately 1.5 m thick grey interval in the otherwise red limestones (e.g. Mellgren & Eriksson 2009; Eriksson et al. 2012). The ‘Täljsten’ likely formed during a rapid regression-transgression cycle and marks a biotic turnover (Eriksson et al. 2012). The interval spans the transition between Asaphus expansus and Asaphus raniceps as well as L. variabilis and Y. crassus, in ascending order (Eriksson et al. 2012). The biostratigraphy and the anomalous facies suggest that the uppermost 0.3 m of the Svartodden Member correlates to the upper part of the ‘Täljsten interval’.

The ‘Täljsten’ is marked by a rapid carbon isotope shift in the Gullhögen quarry of Västergötland (Meidla et al. 2004), a shift which can potentially be traced globally; the base of Y. crassus zone in Maocaopu and Puxi River, China documented by Schmitz et al. (2010), also coincides with a rapid isotope shift. A similar development can be seen in the interval between approx. -17 and -19 m in the Brunflo core, where the thicknesses of the limestone beds are thicker than both below and above and which is marked by a rapid shift in the carbon isotope curve. This shift is rapid and of minor magnitude, but both the decreasing and increasing trends are recorded by several samples and is therefore not an artefact. For these reasons the interval is interpreted as coeval with the ‘Täljsten’. A similar development is seen between -28 and -30 m in the Tingsskullen core, at the top of Formation A+B.
9 Conclusions

The Tremadocian-Darriwilian outcrops of Bjerkåsholmen and Djupetrekodden peninsulas near Slemmestad, Oslo and the Swedish Geological Survey’s Brunflo #2 core were studied with respect to sedimentology and carbon isotope stratigraphy. These were compared to the core from Tingskullen, Öland described by Calner et al. (2014) and Wu et al. (submitted).

The correlation implies a similar sedimentation pattern throughout the western Baltoscandian basin during the studied interval, but the strata are more considerably more expanded in Tremadocian and Floian times in the Oslo area as compared to Jämtland and Öland. The palaeodepth was greatest in the Slemmestad area and most shallow in the Tingskullen core area. Brunflo represent an intermediate position.

The strata are, as a whole, extremely condensed, with average sedimentation rates of only a few millimetres per thousand years. Glauconite is an authigenic mineral forming at low sedimentation rates. The two beds of the Ceratopyge Limestone in Brunflo yield abundant glauconite, and the study shows that glaucony of the upper bed is more evolved, indicating a longer interval of non-deposition than for the lower bed.

Combining microfacies analysis and carbon isotope stratigraphy, several surfaces and intervals important for regional correlation have tentatively been identified for the first time in the Oslo-Asker area and at Brunflo. These include ‘Blomriuga bladet’, ‘Blödläget’ as well as the regressive facies associated with the Volkhow-Kunda boundary and the subsequent ‘Täljsten’ interval. As the biostratigraphic control is stronger in the Oslo outcrop, the degree of confidence in the identifications is stronger here than within the Brunflo core.

Carbon isotope signatures have a great potential for both intrabasinal and global correlations. Several important carbon isotope excursions and trends of the Lower and Middle Ordovician can be recognised in the high resolution carbon isotope data from the Brunflo core. These include the LTNICE in the Latorp Limestone, the Floian-Darriwilian rise in the Latorp and Toyen formations and the BDNIC in the basal part of the Lanna Limestone. A negative excursion is represented in the upper part of the lower member of the Holen Limestone. It immediately precedes the rising limb of the MDICE, which is clearly expressed in the upper member of the Holen and Segerstad limestones.

Caution should be exercised in the study of inorganic carbon isotope signatures for the Ordovician outcrops of the Oslo-Asker area as a significant overprint from the Caledonian Orogeny has disturbed the signal. The organic carbon signature is more stable and provides better data. An extensive degree of recrystallisation and/or pseudomorphosis recorded in the Bjerkåsholmen Formation, Latorp Limestone and Hagastrand Member indicate that evaporite conditions may have prevailed in Baltoscandia during the Lower Ordovician despite being positioned on relatively high latitudes. That sign of evaporitic conditions are not being found in the studied Middle Ordovician strata may be linked to climatic cooling (cf. Trotter et al. 2008). The onset of climatic cooling from the earlier prevailing greenhouse conditions may have caused increased biodiversity and the Great Ordovician Biodiversity Event and with that, increased burial of light organic carbon. This is reflected by the Floian-Darriwilian rise and the Middle-Darriwilian carbon isotope excursion, which have both been documented in the present study.

10 Acknowledgements

First and foremost, I would like to express my heartfelt gratitude to my supervisor Mikael Calner and co-supervisor Oliver Lehnert for introducing me to the project and for all the advice and encouragement you have provided me throughout the process of writing this thesis. Rongchang Wu and Niklas Brädenmark, I am grateful for all the help you have given me and all the fruitful discussions we have had. Git Klintvik Ahlberg and Anders Lindahl have been invaluable in the lab, with instruction and preparation of samples. I would also like to thank Leif Johansson, Carl Alwmark and Mats Eriksson for instructions on how to coat samples and operate the SEM. Anders Lindskog is thanked for the inspirational discussions, as are Per Ahlberg and Britta Smångs for help in locating bibliographical rarities and advice on reference management. Without the comradeship of the fellow students at Geocentrum, writing this thesis would not have been half as enjoyable. Lastly, I would like to thank my family for supporting me through all these years of study.
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